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Recent extreme slope failures in glacial environments: effects of thermal perturbation

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Abstract

This paper describes recent exceptional slope failures in high-mountain, glacial environments: The 2002 Kolka-Karmadon rock-ice avalanche in the Caucasus, a series of ice-rock avalanches on Iliamna Volcano, Alaska, the 2005 Mt. Steller rock-ice avalanche in Alaska, and ice and rock avalanches at Monte Rosa, Italy in 2005 and 2007. Deposit volumes range from 10^6 to 10^8 m³ and include rock, ice and snow. Here we focus on thermal aspects of these failures reflecting the involvement of glacier ice and permafrost at all sites, suggesting that thermal perturbations likely contributed to the slope failures. We use surface and troposphere air temperatures, near-surface rock temperatures, satellite thermal data, and recent 2D and 3D thermal modeling studies to document thermal conditions at the landslide sites. We distinguish between thermal perturbations of volcanic-geothermal and climatic origin, and thermal perturbations related to glacier-permafrost interaction. The data and analysis support the view that recent, current and future climatic change increases the likelihood of large slope failures in steep glacierized and permafrost terrain. However, some important aspects of these settings such as the geology and tectonic environment remain poorly understood, making the identification of future sites of large slope instabilities difficult. In view of the potentially large natural disasters that can be caused by such slope failures, improved data and understanding are needed.

1 Introduction

Landslides larger than 10^6 m³ in glacierized high-mountain environments are rarely well documented because they commonly occur in remote settings involving rock, ice, snow individually or in any combination. Most known landslides larger than 10^7 m³ in high mountains are of

Holocene or Pleistocene age (Abele, 1974; Hewitt, 1988; Strom and Korup, 2006; Korup et al., 2007). In recent years a number of exceptional ice-rock avalanches with volumes of 10^6 to 10^8 m³ have occurred in the European Alps, the Caucasus and Alaska. These events have provided an opportunity to investigate the causes of the failures and the mobility of the related avalanches (Haeberli et al., 2004; Huggel et al., 2005; Fischer et al., 2006; Caplan-Auerbach and Huggel, 2007; Huggel et al., 2008).

Glaciers and permafrost in high mountains are sensitive to thermal perturbations such as atmospheric warming. Thawing can reduce the strength of soil and rock, and can destabilize slopes, leading to failure. We describe here recent landslides at four sites in glacial/periglacial environments where thermal perturbations may have contributed to failure: The 2002 Kolka-Karmadon rock-ice avalanche in the Caucasus (1.1×10^8 m³), a series of high-frequency, high-magnitude ice-rock avalanches on Iliamna Volcano, Alaska ($1-3 \times 10^7$ m³), the 2005 Mt. Steller rock-ice avalanche in Alaska (0.5×10^8 m³), and ice and rock avalanches at Monte Rosa, Italy in 2005 and 2007 ($0.3-1.1 \times 10^6$ m³) (Fig. 1). We specifically investigate the surface and subsurface thermal regimes at these sites, supported by data on surface and troposphere air temperatures, near-surface rock temperatures, satellite thermal data, and recent 2D and 3D thermal modeling studies (Noetzli et al., 2007; Huggel et al., 2008). Based on the four examples, we discuss thermal perturbations that can contribute to or trigger large slope failures in areas with glaciers and permafrost. Time scales at which the different thermal perturbations are effective are considered. We recognize three thermal perturbations that can cause slope instability: volcanic and geothermal activity; glacier-permafrost interaction; and climate change. Our focus is not on the geological and tectonic factors but we acknowledge their importance and possibly dominant control on slope instability. Aspects of avalanche propagation and mobility are beyond the scope of this paper even though they are important for the consideration of hazards.

2 Kolka

2.1 The 2002 ice-rock avalanche

The Kazbek massif in the Northern Caucasus is a glacierized volcanic area and peaks up to about 5000 m asl (42°43'N, 44°29'E). Mt. Kazbek, the highest peak, is considered to be a dormant volcano. Geochemical observations and analyses suggest a magmatic reservoir is present at shallow depth (Polyak et al., 2000). Predominant rock types are lavas, pyroclasts, and metamorphic rocks.

On September 20, 2002, part of the north-northeast face of Dzhimarai-khokh (4780 m asl), a few kilometres west of Mt. Kazbek, failed along a plane at a depth of 30-40 m (Fig. 2; Huggel et al., 2005a). About $10 - 14 \times 10^6 \text{ m}^3$ of rock and about $8.5 - 13 \times 10^6 \text{ m}^3$ of ice detached at 4300 m asl and struck Kolka Glacier at the base of the mountain wall (3000 and 3300 m asl.). Kolka Glacier was almost completely removed by the impact of the slide and about $80 \times 10^6 \text{ m}^3$ of ice entrained into the debris, thus multiplying the avalanche volume. Based on field studies and theoretical considerations, Kolka Glacier appears to have been in unstable conditions at the time of the avalanche, mainly due to load by debris from precursory rock/ice avalanches and quite likely also due to high subglacial water pressure (Haeberli et al., 2004; Kotlyakov et al., 2004; Huggel et al., 2005a). It was thus at risk to be entrained by a high-energy impact of the landslide. The avalanche travelled 20 km down the Genaldon valley at speeds of up to 70-90 m/s. More material was incorporated along the path; at the time it came to rest it increased in size to $115 \times 10^6 \text{ m}^3$. A debris flow continued downvalley for another 15 km stopping 4 km before the town of Gisel. In total about 120 people were killed and roads, residential buildings and other infrastructure destroyed.

Several aspects of the Kolka event are extraordinary, among them the extreme acceleration of the avalanche after the entrainment of Kolka Glacier, the high flow velocity, the long travel distance, and the almost complete removal of the glacier. The Kolka avalanche is also among the largest known historical ice avalanche.

2.2 Thermal conditions

During the assessment of the causes that led to the catastrophic avalanche, the question arose as to what degree permafrost and its possible degradation contributed to the slope failure. First estimates were based on data from meteorological stations at Karmadon and Rokski Pass 15-20 km from Kolka Glacier. Based on the mean annual air temperature (MAAT) at these stations and a lapse rate of $0.006^\circ\text{C m}^{-1}$, Haeberli et al. (2003) estimated an MAAT of $-6 \pm 2^\circ\text{C}$ at the base of the detachment zone and $-11 \pm 3^\circ\text{C}$ at the top. Based on data from the Alps and site-specific aspect, mean annual ground surface temperatures (MAGST) were assumed to be $1-2^\circ\text{C}$ warmer than MAAT (Gruber et al., 2003, 2004). These estimates suggested that the failure zone was frozen at the time of failure.

Subsequently, a more detailed evaluation of thermal conditions was made by installing ten temperature loggers on rock faces in the vicinity of Kolka and Maili glaciers. The loggers were placed on faces with different aspects over an elevation range of 2440 to 3500 m asl at 10 cm depth over periods of more than one year (Fig. 3).

Results show that two loggers at 3250 m and 3500 m asl, with northeast and northwest aspect, respectively, are in permafrost areas with MAGST of about -1.5° to -2.5°C (Fig. 4). One logger at 3000 m asl with northeast aspect is likely at the permafrost limit. The landslide scarp has the same aspect, allowing accurate extrapolation of temperatures from the logger data. The extrapolated values for the base and top of the detachment zone are, respectively, -3° to -3.7°C and -7.5° to -8.5°C . These values are within the range of the estimates by Haeberli et al. (2003) mentioned above.

The calculated MAGST indicate near-surface temperatures for rock under the current climate. To assess their relevance to slope instability, we must consider a number of thermal perturbations that could have affected the failure zone over the past years to decades. The Caucasus has experienced warming for more than a century and temperature have risen more rapidly since the 1970s and in particular since the mid-1990s as reflected in rates of glacier recession (Stokes et al., 2006). On top of climate induced warming effects on rock and firn, steep glaciers on Dzhimarai-khokh are likely to have had a warming effect on underlying bedrock by latent heat dissipation from percolating and refreezing meltwater in the ice. The estimated MAAT lies in a range where generally polythermal or temperate conditions at the ice-bedrock interface behind the cliff exist (Haeberli et al., 2003, see also the section on Mt. Steller). Rising air temperatures can enhance the thermal perturbation caused by steep glaciers on the underlying bedrock by warming the glacier ice. The exposed bedrock is also likely to have been affected by rising temperatures with the warming signal having penetrated several decameters (Noetzli et al., 2007), i.e. to the depth where slope failure occurred in 2002.

An additional thermal perturbation may contribute to destabilizing steep slopes in the Kolka-Kazbek region. The region is tectonically and geothermally active and features a number of geothermally active spots. Heat flow in the Kazbek massif averages about 100 mW/m^2 (Polyak et al., 2000) and several hot springs exist close to Maili and Kolka glaciers and farther downvalley. We have observed surfacing water in the north-facing wall between Dzhimarai-khokh and Maili Glacier. No evidence of geothermal activity has yet been found specifically in the failure zone of Dzhimarai-khokh, but if there was a geothermal flux, either convective or conductive, it would have a strong effect on slope stability.

3 Iliamna

3.1 Frequent large ice-rock avalanches

Iliamna (3053 m asl, 60°02'N, 153°05'W) is an active andesitic stratovolcano located in the Cook Inlet region of Alaska (Figs. 1, 5). The volcanic edifice has developed over Jurassic plutonic rocks (Dettermann and Reed, 1980) and comprises lava flow, lahar, pyroclastic flow and debris avalanche deposits. Major explosive eruptions occurred about 7000 and 4000 years ago (Waythomas and Miller, 1999). Sulfurous fumaroles are present on the east face of the summit, and the volcano has persistent low-magnitude seismic activity. Roman et al. (2004) documented a likely magma intrusion in 1996. The rocks around the summit have been strongly altered and weakened by hydrothermal activity. The amphitheater-like scarps around the summit truncate young lava flows and may have resulted from Holocene sector collapses (Waythomas et al., 2000).

Iliamna is the source of several recent large ice-rock avalanches that have only recently been documented (Caplan-Auerbach et al., 2004; Caplan-Auerbach and Huggel, 2007; Huggel et al., 2007). The earliest documented ones occurred in 1960, and others followed in 1978, 1980 and 1983 (Huggel et al., 2007). Improved seismic, airborne and satellite monitoring of the volcano since the late 1980s has provided evidence of additional landslides in 1994, 1997, 2000, 2003 and 2004. The majority of avalanches documented involved large failures of ice and rock from the east face of Iliamna at elevations of 2000 to 2300 m asl and descended down Red Glacier (Fig. 6). The avalanches ranged in volume from 0.3 to $2.8 \times 10^7 \text{ m}^3$, with six avalanches larger than $1 \times 10^7 \text{ m}^3$ (Huggel et al., 2007). The mode of avalanche failure is unclear because on-site investigation of the summit region is difficult and hazardous. Airborne observations suggest that the avalanches between 1960 and 1983 likely started on sliding planes within glacier ice (Alean, 1984) whereas more recent events involved failure within the volcanic sequence, with failure planes extending into bedrock. Significant amounts of ice and rock were involved in all of the more recent failures. The widths of the scarps range from 600 m to >700 m, and the thickness of ice involved in the failures is estimated to be 20 - 40 m. Avalanche runout lengths ranged from 7.7 km to 10 km. Most of the avalanches entrained large amounts of ice, snow and rock along their flow path, with bulking by a factor of up to 2.

The flow path and dynamics of all major Iliamna avalanches down Red Glacier are similar. For instance, most events resulted in two diverging avalanche flows in the runout and two superelevated flow lobes at the northern margin of the flow path. These features suggest two avalanches separated by a few seconds (Huggel et al., 2007). Flow speeds were estimated at 20-70 m/s using seismic signals produced by the avalanches (Caplan-Auerbach and Huggel, 2007).

3.2 Thermal conditions

Although Iliamna is an active volcano, it has not had erupted in the last 200 years, but features persistent fumarolic activity (Waythomas et al., 2000, Roman et al., 2004). It has been thermally monitored on an irregular schedule by airborne Forward Looking Infrared Radiometer (FLIR), and thermal anomalies related to fumaroles have been detected (McGimsery et al., 1999). More recently, Huggel et al. (2007) used July and August 2003 ASTER night-time thermal infrared (TIR) imagery to study thermal anomalies at the Iliamna summit region. A sub-pixel energy flux calculation to derive the temperature of the heat sources was performed. For assumed sizes of the heat sources from 25 to 1000 m² temperatures ranged from 25°C to a few hundred °C. Evidence of an elevated heat flux at the summit area also comes from airborne observations of liquid water and small debris flows on the southeast, east and northeast flanks of Iliamna (Huggel et al., 2007).

To better understand how the enhanced heat flux may affect slope stability we consider the climatically driven thermal conditions of the glaciers on Iliamna. For steep glaciers with comparably small ice thickness (about 10-50 m), the temperature at the glacier bed can be approximated by the firn temperature, which itself can be estimated using the MAAT (Huggel et al., 2004). We estimated the MAAT at the elevation of the avalanche scarps from the closest meteorological stations to the volcano, at the towns of Iliamna (50 m asl) and Homer (sea level), within 100 km of the volcano. The extrapolated MAAT at an elevation of 2200 to 2500 m asl is -11°C to -15°C, indicating that cold ice conditions should prevail there. Huggel et al. (2004) showed that a relation exists between the ice temperature and the slope of the failure zone, with colder ice being more stable on steeper slopes. Data on ice temperature and slope for ice avalanches in the Alps and on Iliamna by Caplan-Auerbach and Huggel (2007) indicate that Iliamna avalanches occurred on less steep slopes than would be expected considering ice temperatures. This result suggests that ice avalanches at Iliamna could be influenced by a volcanic thermal perturbation, which is not a priori clear due to the spatially inhomogeneous heat flow on volcanoes.

Glacier stability may be affected by permanent or transient geothermal heat. Transported conductively or convectively, increasing geothermal heat flux may cause melt at the base of the glacier. This can result in a reduction of shear strength through increased pore water pressure and reduced effective stress which may lead to failure. Hydrothermally altered rocks in the failure zone beneath the glacier are likely to contribute to the reduction in shear strength, by reducing the cohesion and the friction angle.

An important factor in the high frequency of avalanches at Iliamna may be the high accumulation rate of snow. In the absence of any direct precipitation measurements at Iliamna Volcano, recent glacier mass balance modeling studies in combination with LIDAR derived mass changes on Tuxedni Glacier (Arendt et al., 2002) indicated that snow accumulation in the summit region is probably in excess of 5 m (water equivalent) (Machguth and Huggel, 2008). Such high

accumulation rates at Iliamna would explain the rapid build-up of mass after avalanches and the three-to four year interval between successive avalanches at the same location.

4 Mt. Steller

4.1 The 2005 rock-ice avalanche

One of the largest rock-ice avalanches in the past several decades occurred on Mt. Steller in south-central Alaska (3236 m asl, 60°13'N, 143°05'W) on 14 September 2005. The headscarp of the avalanche is located at 3100 m asl in the glacier-covered south flank of Mt. Steller. The total volume of rock and ice that failed and was entrained along the avalanche path is $40\text{--}60 \times 10^6 \text{ m}^3$. The avalanche traveled 9 km and came to rest on Bering Glacier (Fig. 7; Huggel et al., 2008). Tertiary sedimentary rocks layered sub-parallel to the surface slope characterize the failure zone. Due to its remote location, the avalanche was only detected by seismic signal. The seismic impact of the slide was so strong that it was detected around the world.

The south face of Mt. Steller is about 1600 m high and the failure zone has an average slope of 45°. The upper part of this steep slope is covered by glacier ice (Fig. 8). Airborne observations made a day after the avalanche and a Landsat ETM+ scene recorded just a few hours after the event were used to reconstruct the failure. Huggel et al. (2008) estimated that $3 \text{ to } 4.5 \times 10^6 \text{ m}^3$ of glacier ice failed, based on estimated ice thickness of 20–30 m. Bedrock failed between about 2500 and 3100 m asl; the initial failed rock volume is estimated to be $10\text{--}20 \times 10^6 \text{ m}^3$. Another $2 \times 10^6 \text{ m}^3$ of snow may have been involved in the initial avalanche. Between $5 \text{ and } 30 \times 10^6 \text{ m}^3$ of additional snow, ice and rock were entrained from the south face of Mt. Steller. Examination of aerial photographs shows that several million cubic meters of glacier ice were eroded and entrained from the glacier at the toe of the south face. The avalanche traveled on the surface of that glacier in a laterally confined valley for about 4 km, before reaching the relatively flat surface of Bering Glacier, where the mass spread and stopped with its lowest point at 670 m asl. The seismic signals recorded by stations operated by the Alaska Earthquake Information Center were used to estimate avalanche velocities. Based on the avalanche travel distance of 9 km and the signal duration, the avalanche must have traveled at speeds of up to 100 m/s, possibly even more (Huggel et al., 2008). The high speeds can be explained by the high elevation drop and large avalanche mass resulting in high potential and kinetic energies.

4.2 Thermal conditions

Assessment of the thermal ground conditions at Mt. Steller is a particular challenge because of the remote location, difficult access, and scarcity of ground data. Available climate data was obtained from meteorological stations at Cordova, McCarthy, Chitina, Ernestine and Thompson Pass, which are all less than 200 km from Mt. Steller and range from 10 m to 760 m asl, as well as radiosonde data from Yakutat, 270 km southeast of Mt. Steller. Mt. Steller is located close to the divide of the cold-dry climate of the interior of Alaska and the warmer and humid climate of the coast. We therefore assumed a temperature regime for Mt. Steller intermediate between the coast and the interior. If we rely on extrapolation of the data from climate stations, we derive a MAAT of -14 to -19°C for the failure zone (2500 - 3100 m asl), using a lapse rate of $0.0065^{\circ}\text{C m}^{-1}$ (base on the radiosonde data). The MAGST of the failure zone is estimated to be 2-3°C warmer than MAAT due to its southern aspect, plus another 1-2°C based on rock temperature measurements and comparison to MAAT in the Caucasus (section 2) and in the Alps (Gruber et al. 2003, 2004). These adjustments give a MAGST of about -10 to -15°C for the failure zone. This temperature range is characteristic of cold permafrost where the active layer in summer may be a few decimeters to meters.

Interestingly, the radiosonde data indicate considerably warmer temperatures for the summit region. At the 3000 m asl level an MAAT of -10.5°C is measured (Fig. 9). Due to the relative homogeneity of tropospheric temperatures, the radiosonde data should yield a reasonable estimate even though it was recorded more than 200 km away from Mt. Steller. Radiosonde data result in a MAGST of -6 to -8°C at the failure zone.

At MAAT of about -10°C melting and infiltration processes with latent heat dissipation can start and significantly warm firn and glacier ice. Investigations in the Swiss Alps have shown that steep firn and ice bodies can induce an important thermal anomaly into the underlying bedrock due to such latent heat dissipation from percolating and refreezing meltwater at the firn surface (Haeberli et al. 1997, 1999). Similar situations are typical for glacierized mountain walls and can be found, for instance, at Dzhimarai-khokh (section 2) and Monte Rosa (section 5). Since those first studies, bedrock temperature regimes have been further investigated using 3D thermal models that incorporate surface temperatures (Gruber et al., 2004) and subsurface conduction of heat (Noetzli et al., 2007). Huggel et al. (2008) calculated a 2D temperature profile along a north-south transect through Mt. Steller. Sub-surface temperatures were simulated using a 2D steady-state, finite-element, heat conduction model with a width of 10.5 km and a base at 2000 m below sea level. The upper boundary condition was set by the surface temperature derived from the radiosonde data because those were considered more accurate than estimates from ground-based climate stations due to possible horizontal and vertical extrapolation errors, and inappropriate consideration of

recent warming. The inward heat flux was set at 0.08 W/m^2 for the lower boundary and the thermal conductivity was assumed to be homogeneous and isotropic at 2.5 W/m/K . Homogeneous conditions are obviously a model simplification from natural rock mass characterized by discontinuities that may favor heat advection processes. The focus with these model runs, however, was more on elucidating the effect of steep glaciers on bedrock temperatures. Two modeling runs were performed, one with no ice cover on Mt. Steller, and the other one replicating the situation before the 2005 failure, with hanging glaciers on the Mt. Steller summit ridge. Uncertainties in the assumption of polythermal ice temperature distribution may result in errors of $2\text{-}3^\circ\text{C}$. The model results indicate significant and deep-seated effects of glacier ice on rock temperatures to depths up to several hundred meters in rock (Huggel et al., 2008). Bedrock temperatures in the summit area of Mt. Steller were close to 0°C whereas the area of exposed rock below the hanging glacier in the south face has temperatures clearly below 0°C to a depth of 200-300 m.

In addition to the thermal perturbation by glaciers further consideration should be given to recent rising temperatures, which have been observed in a particularly pronounced manner in Alaska. For the period 1961-1990 warming was $0.015\text{-}0.017^\circ\text{C yr}^{-1}$ whereas it has substantially increased since the mid-1980s to $0.03\text{-}0.04^\circ\text{C yr}^{-1}$ and at some sites in Alaska even more (Chapin et al., 2005; Hinzmann et al., 2005). Warming has been particularly rapid during the past 20 years. Consistent with this warming, the radiosonde data from Yakutat shows a recent warming trend of $0.3\text{-}0.4^\circ\text{C}$ per decade (Fig. 9). Changes in subsurface temperatures are evidenced from long-term borehole observations in low-elevation permafrost areas showing warming rates of $0.05\text{-}0.2^\circ\text{C yr}^{-1}$ in near-surface permafrost since the mid-1980s and a total temperature increase of $1\text{-}4^\circ\text{C}$ (Osterkamp and Romanovsky, 1999; Hinzmann et al., 2005; Osterkamp, 2007). Warming of low-elevation permafrost, however, is less consistent than changes in air temperature because of the influence of local factors such as snow cover and slope aspect (Harris et al., 2003; Osterkamp, 2007). No comparable studies have been done in steep mountain terrain in Alaska. However, limited snow cover on steep rock walls should result in a closer relation between air and bedrock temperatures in low-elevation areas (Gruber and Haeberli, 2007). Hence, 20th century warming in Alaska likely penetrated several meters to decameters into the bedrock of Mt. Steller.

We furthermore found large positive summer temperature anomalies in the radiosonde data. They are evidence of temperatures above freezing persisting at 3000 m altitude as late as September in recent years (Fig. 10). These unusually warm conditions likely enhanced snow and ice melt at the Steller summit. The radiosonde data also showed particularly warm temperatures about 10 days prior to failure (Fig 10), possibly contributing to slope destabilization through increased water infiltration.

5 Monte Rosa

5.1 Ice and rock avalanches

Monte Rosa (4634 m asl; 45°65'N, 7°52'E) is located at the Switzerland-Italy border (Fig. 1). The east face of Monte Rosa is among the highest and most spectacular mountain walls of the Alps, extending from about 2200 to over 4500 m asl (Fig. 11). The face is formed of gneissic and metamorphic rocks and is partly covered by steep glaciers (Fischer et al., 2006). Avalanches from these glaciers feed Belvedere Glacier which extends from the base of the face to near the mountain village of Macugnaga. Starting in 2001, Belvedere Glacier has experienced a surge-like process lifting its surface and increasing in ice velocity up to an order of magnitude (Kääb et al., 2004). In 2002 and 2003 a large supraglacial lake formed at the base of the east face, eventually attaining a volume of $3.5 \times 10^6 \text{ m}^3$, and posing a major flood hazard to Macugnaga (Tamburini et al., 2003).

The steep glaciers on the east face of Monte Rosa have diminished since the end of the Little Ice Age around 1850 (Mazza, 1998), with a dramatic loss of ice during the past 20-30 years (Haeberli et al., 2002; Fischer et al., 2006). Mass movements, mainly rock falls, ice avalanches, and small debris flows, began to increase as the glaciers receded. Evidence of increased slope instabilities includes debris fans that have developed since about 1990 (Fischer et al., 2006). A major ice avalanche from the east face of Monte Rosa occurred in August 2005 (Fig. 11). About $1.1 \times 10^6 \text{ m}^3$ of ice fell from the face but additional snow, ice and debris were entrained along the flow path. Most of the material was deposited in the depression, formerly occupied by the supraglacial lake, but the powder part of the avalanche, including snow, ice and debris fragments overflowed the lateral moraine of Belvedere Glacier and covered the pasture adjacent to an alpine hut. It was fortunate that the avalanche occurred at night, for otherwise tourists who frequently visit the pasture and the alpine hut might have been injured or killed.

Another significant event occurred on Monte Rosa in April 2007 when a mass of rock $\sim 0.3 \times 10^6 \text{ m}^3$ in volume detached from the east face at $\sim 4000 \text{ m asl}$ and fell to its base (Fig. 11). In a similar way to the 2005 avalanche, the rock accumulated in the depression of the former lake. Inspection of the detachment zone by helicopter indicated that the rock mass was layered parallel to the surface slope. A helicopter-mounted LIDAR survey was done in September 2007 to better understand the causes of failures at Monte Rosa (Fischer et al., 2008).

5.2 Thermal conditions

The 2005 ice avalanche initiated on a part of the Monte Rosa east face that experienced particularly strong changes in the past two decades. Slopes immediately below and above the ice avalanche starting zone became completely deglaciated during that time (Fischer et al., 2006). The MAAT at the elevation of the avalanche starting zone is best estimated using data from a climate station at Plateau Rosa/Testa Grigia at 3488 m asl and about 15 km away. The average MAAT at this station for the period 1951-2000 is -5.8°C (Mercalli et al., 2003). The MAAT at this elevation on Monte Rosa should be within of $\pm 2^{\circ}\text{C}$ taking into account possible differences in radiation and other local effects. The steep ice bodies in the east face are hence exposed to atmospheric temperatures of about -5 to -10°C MAAT, which suggests a cold glacier front, but probably polythermal to temperate conditions at some distance behind the front.

The MAGST of exposed rock in the Monte Rosa east face was investigated in more detail by deploying rock temperature loggers at an elevations ranging from 2800 to 3100 m asl and with aspects ranging from east to north. The data indicate that the permafrost limit on north- and northeast-facing slopes is around 3000 m asl (Zraggen, 2005). Results of permafrost modeling studies using rock temperature measurements in the Swiss Alps, alpine climate data and energy balance calculations (Gruber et al., 2004) showed a limit of permafrost at about 3000-3300 m asl (Huggel et al., 2005b). Based on the logger data and the permafrost modeling, we conclude that the east face of Monte Rosa between 3000 and 3500 m asl is a zone of warm and probably degrading permafrost. This zone corresponds to the area of greatest slope instability. Rock slope instability also appears to be concentrated around transition zones between orthogneiss and paragneiss, possibly reflecting different geotechnical properties of these lithologies (Fischer et al., 2006).

The April 2007 rock avalanche occurred early in the year, before summer. However, we note that temperatures during this month set records, with average temperatures $5-6^{\circ}\text{C}$ above the 1961-1990 average in the Swiss Alps (MeteoSwiss, 2007). On some days, melting could have been reached even the elevations of the detachment zone.

6 Synthesis

We propose mechanisms for thermally destabilizing slopes in high mountains that can occur independently but often together: (i) volcanic/geothermal effects; (ii) glacier-permafrost interactions and (iii) climatic change.

6.1. Volcanic/geothermal

Effects of elevated heat flow on ice associated with shallow magmas and interaction between eruption processes and ice have been documented in Iceland, Alaska, the Andes, Kamtchatka, Antarctica and other regions. A considerable body of literature exists on the influence of geothermal and volcanic systems on glacier ice (e.g., Smellie and Chapman, 2002; Clarke and Smellie, 2007), but few of these studies have focused on the stability of ice on steep mountain slopes.

A magma body at shallow depth produces an elevated heat flux near the surface or at the interface between bedrock and a glacier. The surface heat flux depends on the magma temperature (~ 700 to 1300°C), the depth of the magma body (e.g. at Iliamna Volcano 4-6 km; Roman et al., 2004), and heat transport processes. In conductive heat transport regimes, the heat flux at the surface depends on the thermal gradient and the thermal conductivity of the material, which is typically around $2.5 \text{ W m}^{-1} \text{ K}^{-1}$ for solid rock. Heat fluxes of 0.1 to 3.5 W m^{-2} have been measured in geothermal areas where conductive heat transport is important (Whiteford, 1996; Bellani et al., 2004). Heat flux and temperatures at the surface in areas where heat transport is convective can be orders of magnitude higher but varies considerably depending on the settings. Fumarole temperatures may range from a few tens to several hundred degrees Celsius, with heat fluxes of up several thousand W m^{-2} (e.g., Harris and Stevenson, 1997).

Quite obviously, the effects of geothermal heat on ice and slope stability depends on the heat fluxes. A flux of 0.1 W m^{-2} flux over one year melts ca. 1 cm of ice. High-temperature fumaroles of several hundred degrees Celsius can melt tens of meters of ice at the base of a glacier around the vent. In purely conductive heat transport systems, a heat flux of $1\text{-}3 \text{ W m}^{-2}$ which would be expected with shallow magmas could destabilize steep glaciers by melting ice at the glacier base, thus reducing its shear strength. This process possibly occurs at Iliamna (Huggel et al., 2007).

Pressurized hydrothermal fluids are another important hydrothermal perturbation. Hot acidic fluids may alter volcanic rock, producing clay minerals that can reduce the coefficient of friction. Day (1996) proposed that hydrothermal fluids can destabilize the slopes of volcanoes by heating confined pore water and through degassing and pressurization effects (see also McGuire, 1996; Reid et al., 2001). At Iliamna volcano, the avalanche failure zone coincides with a large area of hydrothermally altered rock (Fig. 6) and likely is implicated in frequent failures.

High-temperature fumaroles on volcanoes may break through glacier ice (a few tens of meters thick) (e.g., during eruptions of Fourpeaked Volcano, Alaska, in 2006; and Nevado del Huila Volcano, Colombia, in 2007). Low-temperature fumaroles and conductive heat flow can enhance glacier melt. Because steep glaciers are already at the critical threshold of stability, this melting may

be sufficient to cause a landslide. Although studies on the influence of enhanced geothermal heat flow on slope stability in permafrost environments are missing, steep perennially frozen rock with poor strength will apparently be affected by high heat flow. The effect, although much greater, is spatially much more limited than climatically induced changes in permafrost (see 6.3).

6.2 Glacier-permafrost interaction

The thermal effects of glaciers can be significant in destabilizing permafrost-affected steep slopes at high elevations. The dynamics of glaciers on steep rock slopes depend on both climatic and non-climatic factors, with changes that may be effective within the order of years but are poorly understood in detail (Pralong and Funk, 2005; Fischer et al., 2006). Fischer et al. (2008), for instance, found that the glacier that failed catastrophically on Monte Rosa in 2005 partly regenerated within only two years.

When MAAT is warmer than about -10°C water produced episodically by melting percolates down through firn and ice and refreezes. The freezing releases latent heat, causing the temperature of the ice to increase, possibly to the melting point. At the glacier front, the ice is exposed to atmospheric temperatures and hence is cold and frozen to the ground. Polythermal or even temperate ice behind the glacier front introduces an important thermal anomaly into the underlying bedrock. Such thermal perturbations can increase the rock temperature by several degrees Celsius. Figure 12 represents a model situation similar to that observed at Mt. Steller with a ridge topography with a summit glacier extending into the north and south flanks. The 2D thermal modeling mimicking conditions with and without overlying glaciers reveals the strong thermal effect of glaciers on bedrock temperatures.

At Mt. Steller, liquid water was observed on exposed bedrock in cloudy weather conditions only a few hours after the avalanche at the site where the steep part of the summit glacier failed (Fig. 8). This points to melting at the glacier bed which is corroborated by the high positive temperature anomaly recorded during many days prior to the avalanche (Fig. 10). Rock fractures quasi-parallel to the slope could have facilitated infiltration of water into bedrock. Hydrostatic and effective pressure variations resulting from infiltration and subsequent freezing and volume expansion in fractures could furthermore have caused micro fractures and progressive failure (Eberhardt et al., 2004) in the lower part of the failure zone of Mt. Steller, and thus provide some explanation for failure in the lower part that was not covered by glacier and characterized by probably much colder rock temperatures. However, these considerations remain speculative at the moment, and it is difficult to attribute thermal perturbations due to polythermal glacier to an

observed slope failure because other factors, such as the permeability, structure, and strength of the rock, can have equally important, or even more important effects.

6.3 Climate change

Atmospheric warming during the past decades has been particularly pronounced in mountain and arctic regions, where MAAT has increased up to 0.04°C per year, and even more at some arctic sites (Vuille et al., 2003; Hinzman et al., 2005; Beniston, 2006). For permafrost, long-term monitoring in Alaska and Siberia has shown warming of 0.05-0.2°C per year since the mid-1980s (Osterkamp, 2007; Romanovsky et al., 2007). During about the same period, a ground temperature increase of 0.6°C (at 11 m depth) and secular warming tendencies of 0.5-1°C have occurred in European mountain permafrost beneath slopes of less than about 20° (Harris et al., 2003). In flat or gently inclined terrain, however, snow cover may exert a more important control on permafrost temperature than the atmosphere (Osterkamp, 2007). In steep mountain terrain, where little or no snow accumulates, changes in air temperatures more closely mimic ground temperatures near the surface (Gruber and Haeberli, 2007). However, because temperature monitoring on steep mountain walls has only recently been initiated (Gruber et al., 2004), no corresponding temperature trends are currently available.

Records of ice temperature at high elevations in mountains are sparse. Suter (2002) reported a temperature increase of about 0.5°C throughout a 120 m deep borehole in a glacier near Monte Rosa at 4250 to 4450 m asl between 1983 and 2000. The ice temperatures in that borehole were -12 to -14°C, depending on depth. At a lower-elevation site, where the ice temperatures at depths greater than 20 m were around -10°C, the temperature increased 5-6°C between 1991 and 1999. The difference must lie in an extended period of summer melt at the lower site, resulting in enhanced melt-water infiltration and release of latent heat by refreezing (Suter, 2002). Similar results were recently found on a high-elevation site on Mont Blanc (Vincent et al., 2007).

These findings are significant because they show that the response of glacier ice to atmospheric warming can be strongly non-linear, i.e., the increase in air temperature is amplified in glacier ice. The critical point is when surface melting of ice starts to occur (around -10°C MAAT) or if melting is enhanced as a result of climate change. Warming by release of latent heat upon refreezing can rapidly penetrate to ice depths characteristic of steep glaciers (10-40 m) and can transform initially cold ice into polythermal or even temperate ice within years. Atmospheric warming is thus likely to

be most effective in enhancing and extending the period of melt, producing warmer ice and reducing the friction at the glacier bed.

We are far from having sufficient measurements to document this process in critical topographic situations. Geology will in most cases exert a dominant control on slope stability. Relating observed slope instabilities with warming firn and ice in steep glaciers thus remains speculative. However, there is good reason to think that in the cases of Dzhimarai-khokh, Monte Rosa and Mt. Steller, and possibly Iliamna, warming ice temperatures could have contributed to the failures. In the Caucasus and on Monte Rosa, and based on the radiosonde data also on Mt. Steller, the MAAT was in the range of -5 to -10°C, and firn and ice temperatures of thin ice bodies should be about the same (Huggel et al., 2004). The glaciers at these sites are thus likely to experience infiltration and melt energy input. Extended melt periods probably enhanced this process. In consideration of the temperature difference between extrapolation from climate stations and more recent radiosonde data for Mt. Steller, the recent strong Alaskan warming could have caused summit glaciers to surpass a critical threshold of summer melt with an according non-linear, enhanced response of ice temperatures. The radiosonde data actually document these prolonged melting conditions (Fig 10).

Let us consider a steep glacier with a front of cold ice but temperate ice at some distance behind the front. If an ice avalanche beheads the cold front, possibly due to strength overcome by increased stresses related to warmer ice temperatures and more melt water, then a steep, polythermal to temperate ice body is left from which additional ice avalanches would occur. This scenario likely played out after the large 2005 ice avalanche on Monte Rosa. Bedrock, on the other hand, is suddenly exposed to atmospheric temperatures. Under conditions in which glaciers can cling to steep surfaces, the MAAT is typically well below 0°C. Hence, freezing will penetrate the freshly exposed rock (Kneisel, 2004). When first exposed, unfrozen rock could become unstable. Subsequent penetration of the freezing front could induce stress changes in the rock, for example by volume expansion or ice segregation due to freezing of liquid water. The effects of this rapid thermal perturbation on slopes have not yet been well documented, but it is likely that destabilizing processes are involved.

Knowledge of the temperature distribution and dynamics at depth, and related 3D-effects is important to improve our understanding how climate change affects slope stability. The distribution of permafrost in steep mountain terrain, however, is poorly known. We provide here only a brief overview of some important aspects that are relevant to our case studies. Recently, first attempts have been made to model the temperature and distribution of permafrost in idealized 3D-topography (Noetzli et al., 2007). The current permafrost temperatures at depth are significantly

influenced by the climate of the past millennia, including the last Ice Age. In the perspective of such timescales of heat diffusion, the 20th century warming may only have reached some decameters depth on high steep slopes. Due to the large time lag of heat diffusion, permafrost at greater depth may be present where surface temperatures do not favor it (Noetzli et al., 2007). As a consequence, potential effects on slope stability by recent warming may currently have penetrated to depths of decameters but will continue to reach increasingly greater depths with future warming. An exception may be convective and advective heat diffusion processes that can penetrate much faster and may be particularly favored by rock discontinuity systems (Gruber and Haeberli, 2007). Figure 13 provides a model example of the effect of warming on permafrost degradation in a cross-cut through a mountain.

Although 3D thermal modeling of climate change effects on mountain is an important step forward, it is not yet well understood through which processes such thermal perturbations affect slope stability. Gruber and Haeberli (2007) provided a review on the current knowledge of physical processes likely to be effective in frozen rock with a view on slope destabilization. They also emphasized that thaw is faster and deeper in topographic situations where subsurface heat flow is strongly three-dimensional, such as peaks, ridges and spurs. The recent occurrence of several large slope failures in ridge topography suggests that we have here a good link from theory and model to reality. Mt. Steller is an example of such a south-facing ridge failure, others were observed in the Alps in recent years, i.e. the $2.5 \times 10^6 \text{ m}^3$ rock avalanche from Punta Thurwieser, Italy (3658 m asl), in 2004 (Cola, 2005); and the $2\text{-}3 \times 10^6 \text{ m}^3$ Brenva rock avalanche, Mont Blanc region, Italy (3725 m asl), in 1997 (Barla et al., 2000). Such case studies could also improve the translation of the understanding of thermally related physical processes to slope stability mechanics, which is poorly developed at the moment.

A synthesis of the different thermal perturbations with indicated potential slope failures is shown in Figure 14, with an emphasis on the magnitude of heat injected into the mountain systems by different heat sources. Volcanic heat sources have the potential to be by far the most vigorous thermal perturbation with $10^{-1}\text{-}10^2 \text{ Wm}^{-2}$ by conductive and up to 10^3 Wm^{-2} by convective heat transport. Convective heat transport processes are spatially usually more constrained than conductive ones. Solar and atmospheric short-wave and long-wave radiation with a few hundred Wm^{-2} is spatially unconfined and effective over long periods of time but dependent on the geographic latitude and aspect. It can have important amplification effects in ice as outlined above. The thermal perturbation of permafrost due to overlying glacier ice has a relation to the atmospheric heat source, and varies accordingly. For instance, glacier ice that is at 0°C at the bedrock-ice

interface instead of -5°C theoretically induces an excess energy flux of $\sim 23 \text{ Wm}^{-2}$ although in reality this flux will be lower due to the depth over which the thermal gradients will be effective.

Time scales at which the different thermal perturbations are effective are indicated in Figure 14 and show that volcanic/geothermal heating can vary over shorter periods of time (up to days) than climate-induced warming (years to decades). Heat diffusion of climate warming into the subsurface is generally by conductive processes and usually delayed due to relatively small thermal gradients. Advective heat transport in jointed rock and by infiltration in ice can accelerate and amplify heat penetration into the ground. In volcanic systems heat diffusion can be particularly rapid by convective processes.

7. Conclusions and perspectives

Slope failures larger than 10^7 m^3 are rare in steep high-mountain terrain and very few are known in the 20th century, excluded those triggered by earthquakes or volcanic eruptions. The Mt. Steller and Kolka avalanches with 0.5 to $>1 \times 10^8 \text{ m}^3$ are particularly remarkable in this context. The impact of such large landslides in populated mountain areas can be most severe. Improved monitoring and understanding of how such failures occur are therefore essential to prevent related disasters.

An analysis of air, ice and rock temperatures at four failure sites in three mountain ranges (Caucasus, European Alps, Alaska) suggest that three types of thermal perturbation can be important in destabilizing steep slopes in glacierized high-mountain environments: volcanic/geothermal; glacier-permafrost interaction; and climatically induced warming. Each of the three processes can act separately, but they commonly act together under current atmospheric warming. Time scales at which these thermal perturbations can destabilize slopes are decades to century for conductive heat flow processes, and years to decades for advective/convective heat flow processes. Volcanic thermal perturbation can be effective at even shorter time scales.

It cannot be conclusively demonstrated that slope failures reported in this paper are related to recent warming due to a lack of detailed and long-term data. Similarly, the possibility that these failures are a herald of more frequent catastrophic events in the future is speculative at the moment. The Mt. Steller, Kolka and Monte Rosa cases, however, are potentially linked to atmospheric warming.

The physical mechanisms through which thermal perturbations affect slope stability in cryospheric systems typically result in a reduction of strength of the material. The essential physical processes are not yet understood in sufficient detail but some important mechanisms have been

outlined in this and other recent studies. Future research should therefore try to close the currently prevailing gap between glacier and permafrost thermal models and slope stability mechanics to improve our understanding of current and future landslide hazards in high-mountain environments.

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Figures

Figure 1. Locations of the ice-rock avalanches described in this paper.

Figure 2. View of the upper part of the 2002 Kolka avalanche track. The dashed line demarcates the initial slope failure on the north-northeast face of Dzhimarai-khokh. Also shown is the location of Kolka Glacier before the avalanche, and the path of the avalanche (solid lines) (photo by I. Galushkin).

Figure 3. Locations of the rock temperature loggers around Kolka Glacier; squares and triangles indicate, respectively, permafrost and non-permafrost sites. The failure zone is marked by a dashed line.

Figure 4. Temperature records of two loggers at 10 cm rock depth from January 2004 to July 2005. The aspect and slope inclination of the loggers and mean annual ground surface temperatures (MAGST) are shown for the period January 2004 to January 2005.

Figure 5. Landsat-ETM+ satellite image collected on August 9, 2003, showing runout and deposits of the July 25, 2003, ice-rock avalanche. Older similar avalanches on Red, Umbrella and Lateral glaciers are also indicated. Note the proximity of the fumarole and areas of hydrothermally altered rock to the avalanche initiation area.

Figure 6. Iliamna Volcano showing the path of the July 25, 2003, ice-rock avalanche that travelled down Red Glacier. The failure zone is indicated by a dashed line (photo taken by R. Wessels on August 1, 2003).

Figure 7. Mt. Steller and the rock-ice avalanche track.

Figure 8. The upper part of the Mt. Steller failure zone, a few hours after the avalanche, showing the disrupted summit glacier. Note the flow of liquid water on the exposed rock surface (photo taken by R. Homberger).

Figure 9. Temperature record of the Yakutat radiosonde, 270 km from Mt. Steller, for the period 1994-2007. The altitude of the radiosonde record is approximately the same as the elevation of the Mt. Steller summit. Short-term temperature fluctuation are represented by a 30-day running-

average; the grey curve shows a one-year running average; the straight line is a regression line based on a 30-day running-average, indicating a warming trend of $0.03\text{--}0.04^{\circ}\text{Cyr}^{-1}$.

Figure 10. July-September (JAS) temperature record of the Yakutat radiosonde at 3000 m asl for three years with particularly warm summers. The grey curve represents the 1994-2007 JAS average.

Figure 11. Overview of the Monte Rosa east face. The initiation zones of the 2005 and 2007 avalanches, and the approximate limit of permafrost based, respectively, on corresponding modeling studies by Fischer et al. (2006) and Huggel et al., (2005b) are shown (photo taken by L. Fischer, September 2004).

Figure 12. Subsurface temperature model with conditions similar to those at Mt. Steller. (1) shows the situation without glaciers, whereas (2) indicates the large effect of glacier cover on rock temperatures at depth. Except for the glaciers, all other boundary conditions of the thermal model are the same for (1) and (2) (the base of the glacier on the summit and in the south flank is assumed to be temperate in this example; modified after Huggel et al., 2008).

Figure 13. Subsurface temperature model for steady state (A) and after 100 years (B), simulating heat diffusion effects of a surface warming of $+2.5^{\circ}\text{C}$ and $+3.5^{\circ}\text{C}$, respectively, for the South and North slopes (modified from Noetzli et al., 2007). Note the tilted 0°C isotherm in B, which leads to permafrost persisting at shallow depth but not at the surface of the South flank.

Figure 14. An integrated perspective on different thermal perturbations in a high-mountain system. Magnitudes of heat fluxes and relevant time scales are indicated. Also shown are potential failure planes in rock and ice, as they could be related to infiltration of liquid water.